Estimating Atmospheric Boundary Layer Depth Using COSMIC Radio Occultation Data

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ABSTRACT

This study presents an algorithm for estimating atmospheric boundary layer (ABL) depth from Global Positioning System (GPS) radio occultation (RO) data. The algorithm is applied to the Constellation Observing System for Meteorology, Ionosphere, and Climate (COSMIC) RO data and validated using high-resolution radiosonde data from the island of St. Helena (16.0°S, 5.7°W), tropical (30°S–30°N) radiosondes collocated with RO, and European Centre for Medium-Range Weather Forecasts (ECMWF) high-resolution global analyses. Spatial and temporal variations of the ABL depth obtained from COSMIC RO data for a 1-yr period over tropical and subtropical oceans are analyzed. The results demonstrate the capability of RO data to resolve geographical and seasonal variations of ABL height. The spatial patterns of the variations are consistent with those derived from ECMWF global analysis. However, the ABL heights derived from ECMWF global analysis, on average, are negatively biased against those estimated from COSMIC GPS RO data. These results indicate that GPS RO data can provide useful information on ABL height, which is an important parameter for weather and climate studies.

1. Introduction

The atmospheric boundary layer (ABL) is the conduit through which water vapor and other scalars that are emitted from or absorbed by the surface are exchanged with the free troposphere and thus directly impact weather and climate (Garratt 1994). A distinguishing characteristic of the ABL is turbulent mixing, which minimizes vertical gradients in the mixing ratio of conserved scalars, in contrast to the overlying stably stratified free troposphere, which is characterized by patchy and intermittent turbulence and can support large scalar gradients. A characteristic feature of the ABL top is a substantial decrease in absolute and relative humidity across the top. The depth of the ABL $h$ is determined by a balance between the processes that generate turbulence—vertical wind shear, surface buoyancy flux, and radiative cooling from cloud tops—and mean vertical motion associated with weather systems and large-scale circulations (Medeiros et al. 2005). The extent and types of low-level clouds, and their temporal evolution...
are tied to $h$; however, there are few routine observations of $h$, mostly because it is difficult to systematically measure it, especially over the oceans.

The Global Positioning System (GPS) radio occultation (RO) technique measures parameters of radio signals propagating through the atmosphere between GPS satellites and GPS receivers on low earth-orbiting (LEO) satellites (Rocken et al. 1997; Kursinski et al. 1997; Steiner et al. 1999; Kuo et al. 2004). The phase and amplitude measured as the ray between GPS and LEO ascends or descends through the atmosphere are first inverted to the ray bending angle and next into the vertical refractivity profile under the assumption of local spherical symmetry at the ray tangent point. The retrieved refractivity is a function of pressure, temperature, and humidity. The GPS RO limb sounding technique has been shown to have sufficient accuracy, vertical resolution, and global coverage to have great potential for weather prediction models and climate studies (Healy et al. 2005; Poli et al. 2008; Ringer and Healy 2008; Cucurull and Derber 2008; Cucurull et al. 2008; von Engeln et al. 2009; Kuo et al. 2009; Ho et al. 2009).

Because of the limb viewing geometry, GPS RO is sensitive to vertical gradients of refractivity. The first approach to determine the ABL top from RO was based on truncation of retrieved profiles (von Engeln et al. 2005), which essentially depends on the loss of signal by a generic GPS receiver due to strong defocusing at the sharp top of ABL. Because the refractivity in a typical moist troposphere is mostly dependent on humidity, it is the lapse of humidity across the ABL top that is typically detected in the GPS RO signals and retrieved profiles. Ao et al. (2008) use specific humidity derived from the RO-retrieved refractivity and National Centers for Environmental Prediction (NCEP) temperature for determining ABL depth. Since the humidity is a derived variable, its retrieval requires the use of ancillary information (e.g., temperature). This introduces uncertainties. On the other hand, the lapse of humidity results in a significant lapse of the bending angle and break point in the refractivity profile, which can be used for determining ABL depth without explicit retrieval of humidity (Sokolovskiy et al. 2006, 2007). Basha and Ratnam (2009) found good agreement between the ABL depths estimated from gradients of the RO refractivity and the refractivity, potential temperature, virtual potential temperature, and relative humidity from radiosonde data. They concluded that the use of refractivity for estimating ABL top is superior to the traditional radiosonde methods. Ratnam and Basha (2010) used the RO bending angles and refractivity for estimating the global distribution of the ABL depth and compared them to radiosonde estimates. Their approach, based on the wavelet transform, is equivalent to averaging the profile with a sliding window of fixed length, calculating the lapse within the window, and finding the height of maximum lapse. This is similar to the approach by Sokolovskiy et al. (2007) where the window length was not fixed but was optimally determined for each occultation.

The six-satellite Constellation Observing System for Meteorology, Ionosphere, and Climate (COSMIC)/Formosa Satellite 3 (FORMOSAT-3), launched on 15 April 2006, provides 1500–2000 GPS radio occultation soundings per day (Schreiner et al. 2007). One distinctive feature of COSMIC is open-loop tracking of the L1 (1.575 42 GHz) GPS signal in the troposphere (Sokolovskiy 2001; Sokolovskiy et al. 2009; Ao et al. 2009). This allows measurement of retrieved profiles substantially closer to the surface than was possible with the use of phase-locked loop tracking (Anthes et al. 2008) and provides an opportunity for detailed and extensive studies of the ABL depth.

In this study, we develop an algorithm that can detect the depth of ABLs that have a sharp transition across the top from GPS RO refractivity profiles, and validate this algorithm using radiosonde data from the island of St. Helena (16.0°S, 5.7°W). Based on more than 1 yr of COSMIC RO data, we then study some aspects of the spatial and temporal variability of $h$ and compare the results with high-resolution global analyses from the European Centre for Medium-Range Weather Forecasts (ECMWF). All COSMIC RO refractivity profiles were obtained from the University Corporation for Atmospheric Research (UCAR) COSMIC Data Analysis and Archive Center (CDAAC; http://cosmic-io.cosmic.ucar.edu/cdaac/index.html).

This paper is structured as follows: section 2 presents an algorithm for detecting $h$. Section 3 compares $h$ estimated from COSMIC and radiosonde data. Section 4 presents a 1-yr climatology of $h$. In section 5 we compare $h$ from COSMIC and ECMWF global analyses. Section 6 focuses on the variability of $h$ over the subtropical Pacific. Finally, a brief summary and discussion of future plans are presented in section 7.

2. An algorithm for automated detection of the ABL depth

The highest level product retrieved from the GPS RO signals without the use of ancillary data is the vertical refractivity profile (Kursinski et al. 1997). The refractivity $N$ is related to pressure $P$, temperature $T$, and partial pressure of water vapor $P_w$ by the relation (Bean and Dutton 1968)

$$N = 77.6 \frac{P}{T} + 3.77 \times 10^5 \frac{P_w}{T^2},$$

(1)

where $P$ and $P_w$ are in hectopascals and $T$ is in kelvins. With typical water vapor concentrations in the
troposphere, the sensitivity of $N$ to vertical gradients of $P_w$ is much stronger than to vertical gradients of $P$ and $T$. Further retrieval of the meteorological parameters from refractivity in the presence of water vapor is an under-determined problem and needs ancillary data (commonly, variational assimilation with the use of the first-guess field of an atmospheric model is applied). The purpose of this study is to demonstrate the direct use of $N$ for detecting ABL top without the use of ancillary data.

Figure 1 shows an example of the vertical profiles of $T$ (Fig. 1a), $P_w$ (Fig. 1b), and $N$ (Fig. 1c) with a sharp ABL top at about 1.4-km height (the profiles are based on ECMWF analysis extrapolated in space and time to one of the COSMIC occultations). The ABL top is characterized by a temperature inversion, and a significant decrease of moisture.

As seen in Fig. 1, a significant lapse of humidity across the ABL top results in a break point in the $N$ profile, which can be used for detection of the sharp ABL top (Sokolovskiy et al. 2006). Figure 1c shows two $N$ profiles: the true one (thick line) and the profile that would be retrieved by GPS RO (thin line). The retrieved $N$ profile was calculated by forward modeling of the bending angle from the true $N$ and then inverting it back into refractivity (Abel inversion). The difference between the true and retrieved $N$ profiles is related to the fact that the $N$ gradient at the ABL top in this example exceeds the critical value $-157$ km$^{-1}$ (superrefraction). When this occurs, the retrieved $N$ profile is negatively biased below the top point of critical gradient (Sokolovskiy 2003; Xie et al. 2006; Ao 2007; Xie et al. 2010). Nevertheless, the break point is present in both the true and the retrieved profiles and thus the negative $N$ bias in the case of superrefraction does not impede detection of a sharp ABL top.

We developed the following algorithm for determining the height of a sharp-topped ABL. First, we resample the RO retrieved vertical refractivity profile $N(z)$ with a constant step (5 m in this study). Next we apply a linear
regression $A z + B$ to $N(z)$ in a sliding window of fixed width. We selected 300 m as the window width; we also tried 500 m, but with no significant difference in the results. However, a larger window limits the minimal ABL depth that can be determined by RO that is already limited because of the way RO-retrieved profiles do not penetrate to the surface. A smaller window increases the noise due to small-scale irregularities in refractivity (retrieved by high-resolution wave optics methods; Gorbunov 2002; Jensen et al. 2003) in the lower troposphere. Finally, we define $Z_{bp}$ as the height at which we obtain the maximum lapse of $A$ calculated from the windows immediately above and below $Z_{bp}$, as shown in Fig. 1d (the $N$ profile is the collocated COSMIC RO sounding). For practical use we consider only the RO soundings that penetrate down to at least 0.5 km above mean sea level and satisfy the following criteria: $|A| > 50 \text{ km}^{-1}$ in the window underlying $Z_{bp}$ and $Z_{bp} < 3.5 \text{ km}$. If no ABL top is found below 3.5 km the search is abandoned. We note that the applied criteria are essential even though they limit the number of occultations used for determining the ABL depth. Without the constraint on sharpness the detected ABL top may not always correspond to the actual inversion capping the ABL, and without the constraint on penetration shallow ABLs may be missed and thus the statistical estimates may be biased (Ao et al. 2008; Basha and Ratnam 2009).

3. Validation of the ABL depth detection algorithm

The ABL top is commonly estimated in meteorological profiles by a jump in virtual potential temperature. In the moist troposphere, it is mainly a jump in humidity that affects RO signals. Here, we compare $Z_{bp}$ with the median height $Z_{mw}$ of the maximum lapse of $P_w$, in a 300-m sliding window both calculated from high-resolution radiosonde data over the island of St. Helena (16.0°S, 5.7°W; May–June 2007), where the frequency of occurrence of a well-defined ABL top is high. (Comparison with the maximum lapse of the virtual potential temperature yields similar results.) For consistency with our definition of $Z_{bp}$ we resample radiosonde data with a 5-m vertical step and apply a sliding averaging to $P_w$ with a constant height interval window of 300 m. The results shown in Fig. 2a demonstrate good agreement of the definitions of the ABL top via the $N$ break point and the $P_w$ maximum lapse. The mean difference is related mainly to the fact that, according to the definition, $Z_{bp}$ traces the top of the interfacial layer with maximum lapse of $P_w$.

Next we compare $Z_{bp}$ from COSMIC RO to $Z_{bp}$ estimated in the same way from collocated radiosondes between 30°S and 30°N for May and June 2007. The space and time differences between the COSMIC RO data and the radiosonde data are less than 200 km and 3 h, respectively. The results are shown in Fig. 2b. The standard deviation of the differences between COSMIC $Z_{bp}$ and radiosonde $Z_{bp}$ is about 0.35 km and the mean difference is about 0.04 km, with a correlation coefficient of about 0.82. While the mean difference is quite small, a rather large standard deviation is related to the fact that most of the radiosonde soundings are standard (not high) resolution and also that radiosonde soundings are point measurements while RO soundings represent horizontal averages. This representativeness error may be quite significant since a considerable number of tropical radiosondes are launched from islands that may disturb the ABL and make it more horizontally heterogeneous than over the surrounding ocean.

4. Global analysis of the ABL depth from COSMIC RO data

Here we apply the method for determining ABL depth introduced in section 2 to COSMIC RO data from April 2006 to December 2007. Besides using $Z_{bp}$ to estimate $h$ we also calculate the frequency of occurrence of the ABL with sharp top (hereafter the “occurrence frequency”) as the ratio of the number of RO profiles for which $Z_{bp}$ could be determined to the total number of RO profiles in a given latitude–longitude bin (we use $5° \times 10°$ bins). We note that the occurrence frequency is a subjective characteristic as it depends on the magnitude of the $N$ gradient (averaged over the 300-m height interval) used for selection of the occultations with a sharp ABL top (section 2). However, spatial and temporal variations of the occurrence frequency objectively characterize variations of the sharpness of the ABL top.

Generally, with the applied constraints (discussed in section 2) $Z_{bp}$ estimates are much more frequent over the oceans than over land because of the more homogeneous surface over the ocean and its large thermal inertia, which reduces horizontal height variations in $h$, compared to the ABL over land. We note, however, that by modifying constraints it is possible to search for the ABL depth over certain continental regions (Sokolovskiy et al. 2007). Figures 3a–d show global distributions of $Z_{bp}$ occurrence frequency for four seasons [December–February (DJF), March–May (MAM), June–August (JJA), and September–November (SON)] from April 2006 to December 2007 (JJA and SON include full data from both years while DJF and MAM include full data from one year and partial data from another year). Because of the low frequency of occurrence with the applied constraints, the data over continents are not shown. The
highest frequency is found in the subtropics, with only rare occurrences in the intertropical convergence zone (ITCZ). High occurrence frequency is found throughout the year off the western coasts of continents (North and South America, Africa, and Australia). These regions of high occurrence frequency of $Z_{bp}$ are associated with the persistent stratocumulus and trade cumulus regimes and generally subsiding motion. In the ITCZ, the occurrence frequency is low (the ABL top is ill defined) because intense deep convection transports near-surface air into the overlying free troposphere and smears out the ABL top. Also, the occurrence frequency is low in high latitudes because of small humidity gradients and because the ABL may sometimes be too shallow to be captured by RO. Generally, we believe that estimates of $Z_{bp}$ should be used with care in the regions of low occurrence frequency. When averaging $Z_{bp}$ over regions with substantially nonuniform distributions of samples, the $Z_{bp}$ estimates obtained from RO should be weighted to remove the effects of the nonuniformity (caused by horizontal variability in the sharpness and height of the ABL inversion and nonuniformity of RO sampling).

Figures 3e–h show global distributions of $Z_{bp}$ for the four seasons from April 2006 to December 2007. Deep ABLs are found over the warm pool of the western Pacific, Atlantic, and Indian Oceans, and the ITCZ. Note that $Z_{bp}$ decreases toward the western coasts of continents (North and South America, and Africa) because of the strong subsidence and relatively cold upwelling water. A notable seasonal variation in frequency and magnitude of $Z_{bp}$ is observed over the northwestern Pacific region, which is known for its intensive cyclogenesis (especially from July through October). In that region, the highest $Z_{bp}$ frequency and magnitude occur during the winter (DJF), and the lowest occur during the summer (JJA).

5. Comparison of the ABL depths from COSMIC and ECMWF analyses

Here we compare the ABL depths obtained from the COSMIC RO data with those obtained from the ECMWF global analyses. This is perhaps the most robust method available to assess the validity of the ECMWF estimates of $h$. Since $h$ is the net result of deepening by entrainment and shallowing by subsidence (Medeiros et al. 2005), this is a sensitive indicator of how well the model represents these competing processes. The global ECMWF operational analysis fields of temperature, water vapor, pressure, and geopotential height have previously been studied for the climatology of ducting (von Engeln and Teixeira 2004). The ECMWF analysis fields used here cover NH spring (MAM) to NH summer (JJA) over the year 2007. We note that during this period the ECMWF was assimilating COSMIC data (bending angles) but only down to 4-km height (i.e., mostly above the ABL top over the oceans). The ECMWF grid fields have 91 vertical levels (with 21 levels between 0 and 3.5 km), about 25-km horizontal resolution, and are available at 0000, 0600, 1200, and 1800 UTC. The vertical resolution is roughly 200 m at 1.5-km height. The ECMWF global analysis fields are linearly interpolated to the times and locations of the COSMIC RO profiles. Refractivity profiles are calculated from the ECMWF analysis fields of temperature, water vapor,
FIG. 3. (a)–(d) Global distributions of the occurrence frequency of sharp ABL tops and (e)–(h) $Z_{up}$ for the four seasons from COSMIC RO data in 2006–07.
pressure, and geopotential height. Next, the ECMWF refractivity profiles are log-linearly interpolated onto the height grid of the collocated RO profiles. Finally, $h$ is calculated from the ECMWF interpolated refractivity profiles using the same algorithm as for the COSMIC RO profiles (section 2).

The differences between the ABL depths estimated from COSMIC and ECMWF are additionally affected by the vertical discretization of the ECMWF fields. To evaluate this effect we focus on the subtropical eastern Pacific region where strong inversions are associated with large subsidence and convective ABL mixing due to radiative cooling from the tops of boundary layer clouds. The domain extends from $0^\circ$ to $30^\circ$S and $100^\circ$ to $150^\circ$W from March to August 2007. Owing to the limited ECMWF vertical resolution, there are distinct differences between the ECMWF and COSMIC distributions of $Z_{bp}$ (shown in Fig. 4) due to clustering around the mean heights of the model levels. However, because the spread of $Z_{bp}$ is substantially larger than the difference between ECMWF mean level heights, the clustering does not impede revealing a systematic difference between COSMIC and ECMWF ABL depths.

Figures 5a–d show the occurrence frequencies of $Z_{bp}$ estimated from both COSMIC RO and global ECMWF analyses during NH spring (MAM) and summer (JJA) of 2007. We see that the horizontal distribution of COSMIC $Z_{bp}$ occurrence frequency, although somewhat higher, in general agrees relatively well with the ECMWF analysis.

Figures 5e–h show the ABL depths estimated from COSMIC RO and global ECMWF analyses during NH spring (MAM) and summer (JJA). While COSMIC and radiosonde ABL depths compared in this study are generally in good agreement, as mentioned in section 3, the ABL depth estimated from ECMWF model data, on average, is lower than that from COSMIC data (the difference between the ECMWF and COSMIC ABL depths varies with regions and seasons). Von Engeln and Teixeira (2004) also show that the mean altitude of the ducting layer reproduced by ECMWF has a negative bias of about 150 m in comparison with that from Vaisala RS80 radiosondes. Again, we note that although the refractivity retrieved from RO below the ABL top may be biased because of superrefraction, there are no physical reasons for biases in the ABL depth (except when the very shallow ABL is not captured by RO).

The differences in $Z_{bp}$ between COSMIC and ECMWF suggest that there are some limitations in representation of the ABL depth by the ECMWF model. This is not surprising since $h$ results from a delicate balance between the large-scale subsidence and entrainment, neither of which is modeled well by large-scale models with limited vertical resolution. This is common to all NWP and climate models and is not specific to the ECMWF model (von Engeln and Teixeira 2004). For example,
Fig. 5. Global distributions of the occurrence frequency of (a)–(d) sharp ABL top and (e)–(h) ABL depth during spring (MAM) and summer (JJA) of 2007 for (left) COSMIC and (right) ECMWF. Lines a and b in (a) show the positions of the transects discussed in Fig. 6.
Zeng et al. (2004) found a systematic overestimate of $h$ by CCSM2 when compared with radiosonde soundings from the eastern Pacific subtropical region.

6. The ABL over the tropical Pacific Ocean

The ABL over the cold upwelling ocean on the eastern boundaries of the subtropical Pacific basin (as well as other similar eastern boundary regions) is characterized by a well-mixed, stratocumulus-topped ABL with a well-defined top. Farther from the coast the ABL deepens and the stratocumulus transitions into a trade cumulus regime (Bretherton and Pincus 1995; Wood and Bretherton 2004). The ABL depth in the subtropics is well correlated with the type and amount of cloud cover. Shallower boundary layers are associated with stratiform clouds with larger percentage cloud cover; farther west, the stratocumulus break up and the deepening ABL is associated with trade-wind cumulus and less cloud cover.

The transition in $h$ from the stratocumulus to the cumulus regions in the subtropical Pacific Ocean on both sides of the equator is reflected in the COSMIC $Z_{bp}$. We averaged $Z_{bp}$ in $5^\circ \times 5^\circ$ latitude–longitude bins and plotted its magnitude and frequency of occurrence versus longitude from the eastern Pacific coast line westward to a distance of 45° for all seasons. The results for the period from April 2006 to December 2007 are shown in Fig. 6. The left panels show averaged $Z_{bp}$ (vertical bars show standard deviations); the right panels show the occurrence frequency. The top panels represent the transect 28°N, 115°W to 20°N, 160°W (west of the coast of California peninsula to Hawaii); the bottom panels show the transect 27°S, 75°W to 27°S, 120°W (west of the coast of Chile); both transects are shown by black lines in Fig. 5a. The mean COSMIC ABL depth is below 1.5 km along the California and Chile coasts and increases toward the west up to about 2.3 km near Hawaii (NH) and 120°W (SH).

The occurrence frequency increases from the California coast westward from about 60% to about 95% and then decreases to about 75% near Hawaii. The small values near the California coast may be due to occurrences of very shallow $h$, below the minimum height for which $Z_{bp}$ can be estimated, and the decrease near Hawaii is the result of the deepening of the trade cumulus regime.

![Fig. 6. (left) ABL depths and (right) frequencies of occurrence estimated from COSMIC RO for four seasons in the (top) North and (bottom) South Pacific; see text for details. Vertical bars in the left panels show standard deviations.](image-url)
This increases horizontal heterogeneity in $h$ due to the more intense convection occurring in the scattered and sometimes organized and precipitating patterns of cumulus clouds compared to relatively homogeneous strato-cumulus clouds near the coast (e.g., Painemal et al. 2010), and a thicker transition layer at the ABL top, which scales with $h^{2/3}$ in the clear air ABL according to Gryning and Batchvarova (1994).

There is a significant difference in occurrence frequency between the NH and SH. In the SH, near the Chile coast, the occurrence frequency is close to 1 during all seasons. In the NH, the occurrence frequency decreases toward the western end of the domain. Also, in the SH there is a pronounced seasonal variation of the occurrence frequency at the western end of the domain: minimum during autumn (MAM) and maximum during spring (SON). The seasonal variation of the ABL depth is not well pronounced except somewhat at the eastern part of the SH domain where the deepest and the shallowest ABLs are observed during autumn and winter, respectively.

7. Summary

We developed a method for estimating the ABL depth from RO data. The method uses the break point in the vertical refractivity profile. This coincides with a significant decrease of humidity across the top of the ABL. The ABL depths obtained with this method are in good agreement with those estimated by using lapse in humidity.

The ABL depths obtained with the developed method from COSMIC RO data were validated against the depths obtained from nearby high-resolution radiosonde data and compared with ECMWF global analysis fields. While there is no significant systematic difference between COSMIC RO and radiosonde ABL depths, the ECMWF ABL depths are negatively biased compared to COSMIC RO.

The developed method was applied to COSMIC RO data from April 2006 to December 2007. From this, we constructed global maps of the ABL depth over the oceans for all four seasons to reveal its global and seasonal variations. The method is particularly robust in the subtropical high pressure regions at the eastern ends of the ocean basins because of the well-defined inversion and horizontal homogeneity in these regions. Therefore, we carried out a more detailed analysis of the ABL depth and frequency of occurrence of a sharp ABL top at the eastern end of the Pacific Ocean in the subtropics, which showed, for example, a strongly sloped ABL top that descends from a mean of about 2 km at 150°W to about 1.5 km at 120°W. This analysis gives estimates of the temporal and spatial variability of the ABL depths in these regions, which can be used as a basis for comparison with model predictions.

This first attempt demonstrates the power and usefulness of the method. With more RO data and further refinement of this technique, more detailed analysis of the ABL depth and the frequency of occurrence of a sharp ABL top on different spatial and temporal scales, including diurnal variations and weather-related variations, will be possible.

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